Precambrian Tectonic Evolution of Earth: an Outline

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Abstract

Space probes in our solar system have examined all bodies larger than about 400 km in diameter and shown that Earth is the only silicate planet with extant plate tectonics. Plate tectonics is unusual in our solar system and may be unusual in time. Venus and Earth are about the same size at 12,000 km diameter, and close in density at 5.2 and 5.5 Kg.m-3 respectively. Venus and Mars are stagnant lid planets; Mars may have had plate tectonics and Venus may have had alternating 0.5 Ga periods of stagnant lid punctuated by short periods of plate turnover. Plate tectonics has clearly operated on Earth since the beginning of break-up of Rodinia at about 0.7 Ga, witnessed by rock associations such as obducted supra-subduction zone ophiolites, blueschists, jadeite, ruby, continental thin sediment sheets, continental shelf, edge, and rise assemblages, collisional sutures, and long strike-slip faults with large displacements. Equally, from rock associations and structures, nothing resembling plate tectonics operated prior to about 2.5 Ga. Contentious questions are: “when did plate tectonics start”, “did plate tectonic style and rock assemblages evolve with time?”, and “what tectonic mechanism(s) was responsible for shaping pre-plate tectonic Earth?” The many opinions on these issues have been summarised by Korenaga (2013). We conclude, following Burke and Dewey (1973), that there is no evidence for subduction before about 2.5 Ga, and that plate tectonics or, at least, some form of large lateral relative displacement mobilism evolved during the period from 2.5 to 2.1 Ga after which “modern assemblages”, and long linear/curvilinear deformation belts are developed, and palaeomagnetism indicates that large lateral relative motions among continents had begun since at least 1.88 Ga. Prior to 2.5 Ga there was a stagnant lid. The ”boring billion”, from about 1.8 to 0.8 Ga, was a period of two super-continents, Columbia and Rodinia with substantial intra-plate magmatism and marginal accretionary tectonics.

Modern plate tectonics from about 0.8 Ga is correlated with major glaciations, including the Snowball Earth and the appearance of metazoan life. Our conclusions are based, almost wholly, upon geological data sets, including geochemistry, with minor input from modelling and theory.

Introduction.

Plate tectonics *sensu stricto* (Isacks et al., 1968; Le Pichon, 1968; Mckenzie and Parker, 1967; Morgan, 1968; Wilson, 1965) is the relative motion among torsionally rigid plates with narrow boundary deformation zones in the oceans but, generally, wider in the continents. Significant intra-plate deformation in the oceans is rare, except very close to ridge axes but minor deformation is common in the continents, except for the stiff Archaean cratons with a thick lithosphere, which are almost earthquake-free and commonly wrapped around by younger orogenic terrains. Plate tectonics is a highly-efficient mechanism for global heat-loss by magmatism and hydrothermal convection at the oceanic ridges, with minor conductive heat-loss, supplemented, episodically, by intra-plate magmatism in large igneous provinces and hotspots above mantle plumes, and cooling by subduction. If there was a time on Earth before plate tectonics, neither conduction nor radiation can account for the necessary heat-loss so that the only credible mechanism would have been pervasive plume upwelling and related massive mafic magmatism with, possibly, some form of crustal-lithospheric fragmentation and foundering, drip, sagduction, and localised, non-connected small subduction zones to allow surface materials into the mantle. Mantle temperature and heat-loss have been diminishing since the origin of Earth (Bickle, 1978; Bickle, 1986; Korenaga, 2013); therefore one might expect changes in tectonic style, involving a stagnant non-segmented lithospheric lid to plate tectonics.

Plate tectonics, at a planetary scale, in our solar system, appears to be restricted, now, to Earth but may have occurred in other planets, especially Venus, and may also be occurring in the moons of the giant gas planets. There has been a plethora of suggestions and substantial disagreement on when plate tectonics began on Earth, and what tectonic regime may have applied before plate tectonics, based upon a large range of criteria (e.g. Cawood et al., 2018; Johnson et al., 2017; Moyen and van Hunen, 2012; Smithies et al., 2007). Also, there has been a natural tendency to lean on the doctrine of "the present is the key to the past" perhaps with the notion that plate tectonics *sensu stricto* is too good a mechanism to waste. Indeed, since the advent of plate tectonics (Wilson, 1965), there has been a naturally enthusiastic but over-zealous tendency to interpret the whole recorded history of Earth in these terms. Uniformitarianism has meant a variety of things to geologists. The episodicity, periodicity, and catastrophic geologically-instantaneous nature of many if not most processes indicates that uniformitarianism is not strictly true, though this statement depends upon the time over which processes are differentiated and integrated. It is clear that tectonic processes were not uniform through time and that various forms of secular evolution have occurred, as shown in the compilations by Bradley (2011) of "everything through time". Heat production, the temperature of the asthenosphere, and heatflow have evolved with time, from three times the heat production in the early Archaean compared to today, with a corresponding progressive thickening and stiffening of the lithosphere, whatever form of tectonics is operating.

There is a plethora of opinions, summarised by Korenaga (2013), on the existence, origin and nature of "plate tectonics" before 0.8 Ga. Stern (2005) argued, from plate subductability, blueschists, ruby, jadeite, and ophiolites, that modern plate tectonics began at about 0.7 Ga, with a stagnant lid prior to that, interrupted by a period of plate tectonics from about 2.1 to 1.7 Ga. Hamilton (2007) argued for a 2.0 Ga start based upon the first large deformation belts, Grieve (1980) for 2.7 Ga, Brown (2006) for 2.8 Ga, Polat and Kerrich (1999) before 3.8 Ga, Nutman et al. (2002) at 3.6 Ga, Komiya et al. (1999) and Kusky et al. (2018) at 3.8 Ga, Shirey et al. (2008) at 3.9 Ga, Hopkins et al. (2008) at 4.2, and Harrison et al. (2005) before 4.4 Ga. There is no doubt, from palaeomagnetism (Evans and Pisarevsky, 2008), that substantial relative horizontal motions took place during the Proterozoic from about 1.9 Ga. Smithies et al. (2005a), Cawood et al. (2006), Dewey (2007), Pease et al. (2008), Condie and Kroner (2008), Brown (2008). Van Kranendonk et al. (2007) and Dhuime et al. (2011; 2012; 2015) argued from geology, and numerical modelling, that some form of plate tectonics or, at least, lateral relative motion driven by subduction, started at the beginning of the Neo-Archaean at about 3.1 Ga, marked also by an increase in the growth rate of the continental crust. The uniformitarian plate tectonic view was challenged, initially, by Burke and Dewey (1973), who saw the Archaean as a "permobile" regime characterized by pervasive magmatism and deformation during density inversions when no part of Earth, on any scale, consisted of torsionally rigid lithospheric plates *sensu stricto* on any scale. They argued that plate tectonics began during the early Palaeoproterozoic.

If plate tectonics is the operator, one might expect to see geological evidence of plates and plate boundaries such as the following: passive continental margin and shelf associations; supra-subduction zone ophiolite complexes *sensu stricto* indicating both fore-arc plate accretion and the relative horizontal motion of obduction; long linear and curvilinear deformation belts between widespread, flat-lying, little or undeformed epicontinental platform sequences: transcurrent faults with large displacements; paired metamorphic belts with adjacent high pressure-low temperature and high temperature-low pressure zones, adakites; subduction-accretion prisms; collision zones with ruby; foreland fold thrust belts especially thin-skinned; and palaeomagnetic evidence of the relative motion of continents. These plate tectonic rock suite indicators might be expected to be arranged in belts and zones rather than in blobby patches.

It is interesting to speculate on how we might approach the problem of Archaean tectonics were we to know nothing of plate tectonics. We know that Earth has evolved thermally and so perhaps it should not be surprising that the principal heat loss mechanisms of plumes and plate tectonics have also evolved. Many papers depend on one or a small number of distinctive rock types, such as boninite and andesite, but these are not definitive indicators of subduction because the can also form, even in modern Earth, in intraplate settings. For example, boninites have been found in The Manihiki Plateau (Golowin et al., 2017) and abundant calc-alkaline andesites in the Basin and Range (Hawkesworth et al., 1995). Thus, reliance upon certain rock types and chemistries to give definitive solutions may yield incorrect answers.

This paper focuses on rock assemblages, patterns, and relationships, with accompanying geochemistry and petrology, that define tectonic regimes and involve mostly field observations and geological maps. In simple terms, the question is asked "do geological patterns, at all scales, resemble those that we know were generated by plate tectonics?" or do they differ. Changing heat production suggests the likelihood of progressive rather than instantaneous change, perhaps with tipping points. If we see a geological pattern like, for example, that of the Lower Palaeozoic of western Newfoundland, we may suppose that it was generated by an oceanic arc with an ophiolitic fore-arc colliding with a rifted continental margin. Colloquially, avoiding ducks, "if it looks and behaves like a cat, it probably is a cat". If it has scales and two legs and swims, it probably is not. Insufficient attention has been paid to rock suites and their structures and arrangements that are seen in outcrop and across regions. For example, andesite, tonalite, and diorite are not a definitive indicators of plate tectonics unless they have arc chemistries and are in linear-arcuate belts that, in turn, indicate a related subduction zone. Also, too much reliance has been placed on numerical modelling at the expense of empirical field data; modelling never trumps observation. The key issue is tectonic style and observed geology. Obviously this is not possible for the Hadean for which we have only a handful of primitive zircons, and where modelling and inference are necessary.

Plate tectonics can be dealt with in fully quantitative terms from 160 Ma by sequential magnetic anomaly and continental margin fitting from which poles and rates of relative plate motion can be deduced. Semi-quantitative plate tectonics is allowed by palaeomagnetism, faunas, facies, and paleoclimatology from the beginning of the Cambrian to the early Jurassic because oceanic crust with its magnetic anomalies has been subducted and vanished along suture zones. In the Pre-Cambrian, there are no faunas of use so that palaeomagnetism and facies allow only sketchy continental reconstructions and analyses of relative motion. All those who have written on the subject and, probably most geologists, would agree that plate tectonics *sensu stricto* can be traced back, at least, to the break-up of the supercontinent Rodinia from about 0.8 - 0.6 Ga.

The existence of Rodinia requires that it was assembled from older continental masses and fragments involving subduction and collision. It is commonly considered that this was completed at about 1.0 Ga when the Grenville and Namaqua Orogens were formed (Fig 1). There is clear evidence from palaeomagnetism (Evans and Pisarevsky, 2008) that relative motion of continental masses (continental drift) was occurring back to at least 1.8 Ga and is permissible back to about 3.0 Ga. Extensive linear to arcuate zones of continental deformation, the oldest of which are the Limpopo and Ubendides of southern Africa at about 2.0 Ga (Fig. 2), are characteristic of continental collision zones. Perhaps the finest and clearest example of a Proterozoic collision zone is the Wopmay Orogen (Hoffman and Bowring, 1984), which is part of an extensive network of late Palaeoproterozoic orogens. These include the Trans-Hudson and Labrador Trough, the latter with the characteristic paired positive-negative gravity anomaly, indicating that that the lithosphere was sufficiently strong to support it without relaxing non-elastically to perfect Airy isostasy. Moreover, there is no obvious horizontal strain in the cratons at the time that the bounding Palaeoproterozoic orogens were developed, evidence that they were torsionally-rigid platforms. Care should be exercised with this argument because even modern continental plate boundary zones may be very wide and complex with large torsionally rigid blocks; this is seen in the Himalaya-Tarim and the Cordilleran-Rocky Mountain plate boundary zones. A further problem is that, if oceans existed during the Precambrian, they have been subducted, except possibly in small protected patches like the South Caspian. In Phanerozoic suture zones, ophiolites *sensu stricto* evolved in fore-arcs; therefore, although not ocean floor, an ophiolite fore-arc indicates the subduction of oceanic lithosphere. It is fairly certain, from these general arguments, albeit with caveats, that relative motion between and collision of continents was occurring back to at least 2.0 Ga, with a lithosphere of some torsional strength suggestive of plate tectonics (Fig.1). It can be said, with some certainty, that plate tectonics was operating in its present form since about 0.8 Ga, and something at least resembling plate tectonics since 2.3 Ga. Therefore, in this paper, we explore, mainly, the tectonic history of Earth before 2.0 Ga.

Related critical questions are the volume, composition, and areal distribution of the continental crust. Were there many early small continents or was there a time when continental crust covered the globe with a subsequent loss (Armstrong, 1968; Fyfe, 1978) by massive tectonic erosion leaving stable deformation-resistant cratons? If plate tectonics was not operating in a world of discrete continents, what was between the continents and what was its composition? How can the rate of crustal addition and growth be judged, and how much crustal subtraction and loss took place (Dewey and Windley, 1981)? If there were discrete continents in a stagnant lid, what was happening in the "oceans" between them? What are the lithospheric conditions necessary for subduction? Without subduction, what were the mechanisms for getting water into the mantle to make continents.

Plate tectonic indicators in the rock record

We now look at the temporal distribution of a number of features that are characteristic of Phanerozoic plate tectonics and providing confidence that plate tectonics was operating if many or all were present at a point in space and time where a range of structural, petrological, facies, and stratigraphical features in the geological record were helpful in distinguishing tectonic regimes. Also, there are geochemical trends and characteristics that help to define the evolution of the mantle and continental crust. A particular problem is that there are time gaps in the occurrence of some features, which may be genuine gaps because of episodicity (e.g. four main times of ophiolites during the Phanerozoic) and periodicity, or false gaps caused by non-preservation at the surface by subduction, subduction-erosion or temperature overprint. A good example is ultra-high pressure metamorphism, which is almost entirely younger than 600 million years with outliers at about 2.3 Ga. Also a particular key feature of plate tectonics, such as blueschists, may not have developed in a hotter regime. Times of supercontinents will have an abundance of marginal arcs, much intra-plate rifting and potassic magmatism, and widespread continental sediment sheets. Times of distributed continents will have abundant rifted margins, oceanic arcs with ophiolite fore-arcs colliding with rifted margins, and Andean-style orogens. We should not expect to be able, always, to tick off a convincing list of features that demand a plate tectonic solution. Figure 1, shows some key events and sequences in the tectonic history of Earth.

Hadean (4.55 - 4.0 Ga).

Following the Gaia-Theia collision at 4.51 Ga, the 4.4 Ga Jack Hills zircons in Australia and the Palaeoarchaean 4.03 Ga Acasta gneiss in Canada are the only physical remnants of an event on Earth during the Hadean. The nature of the Jack Hills zircons and the lithology from which they were derived, whether silicic or, mafic, is disputed. Harrison et al. (2005) have argued that the trace chemistry of the zircons points to plate tectonic processes and the growth of continental crust for their origin. We suggest that an opposite conclusion may be drawn from simple geological reasoning. If there had been a substantial amount of Hadean granitic, island arc, or TTG crust, why did none of it survive? The survivability of buoyant, weak continental crust is the reason for mountains and for the preservation of TTG rocks from 4.03 Ga. The lack of extant Hadean continental crust implies that there never was any and that the Hadean crust was largely mafic and later subducted. Furthermore, the absence of granitoids suggests the absence of water until the first TTG's appear in the Eo-Archaean and no plate tectonics.

We envisage the Hadean world to have been, like the Eo-Archaean Moon, one of hot, anhydrous conditions in a stagnant mafic lid, bombarded by bolides. Our reconstruction of the Hadean is similar, in its essentials, to that of Grieve (1980), summarized briefly as follows. Hadean Earth was a dry planet with heat production two to four times that of today (Bickle, 1978; Bickle, 1986; England and Bickle, 1984). The stagnant lid lithosphere must have been thin with a fifteen kilometre basaltic-komatiitic plateau-like crust, shallower than today, generated by a high degree of partial melting of hot mantle (Debaille et al., 2013; Fischer and Gerya, 2016). Continuous bolide bombardment generated multi-ring impact basins with massive excavation, lithospheric fracturing, and mafic volcanism induced below impact sites with both shallow and deep plumes as the mechanisms of global heat-loss and the origin of a probably komatiitic crust. Small felsic pools may have developed, from which the Jack Hills zircons were derived, and there was likely reworking of intra-basin volcanics, and impact lithologies. Impact craters, such as those on the Moon, probably reflect the geometry and topography of Earth's surface in Hadean times, whereas the Vredefort crater in South Africa, although younger at 2025 Ma, preserves the geological responses to cataclysmic impacts upon both basement granitic crust and younger sediments. The only Hadean rocks to have been found to date remain the Acasta Gneiss (Bowring et al., 1989b).

Eo-Archaean (4.0-3.6 Ga)

The main key to understanding the early tectonic behaviour of Earth is the plagioclase-rich tonalite-trondjemite-granodiorite (TTG) of the earliest continental crust, the Ancient Grey Gneiss (AGG), exposed in the Kaapvaal, Greenland, Slave (Acasta Gneiss; Bowring et al., 1989a, b; Bowring and Williams, 1999), Zimbabwe, and Pilbara Cratons. The AGG is, in places, overlain unconformably by Archaean komatiite sequences but, more commonly, has tectonic contacts at dome margins with mafic and ultramafic volcanic and sedimentary sequences (greenstone belts). TTG's vary substantially from low-pressure derivatives, generated by the hydrous partial melting of garnet-free, plagioclase-rich, amphibolites and low-magnesium basalts at up to 900°C at about 35 km, to high-pressure from garnet – amphibolites or eclogites at greater than 60 km depth (Adam et al., 2012). Eoarchaean TTG's are low pressure. Therefore, the crust and mantle must have acquired water by about 4.0 Ga and there must have been either a very thick (>35 km) plateau-like basaltic crust and/or a mechanism to transport mafic rocks into the shallow mantle by subduction or sagduction. We suggest that Earth's water was acquired from icy comets during the late heavy bombardment, which also fractured the lithosphere to allow the sinking of giant mafic slabs that could have been the source from which the TTG's partially melted. We suggest that a thick mafic crust on a thin lithosphere with a geothermal gradient three to four times that of today, perhaps aided by impact-induced foundering, was the likely TTG source. Johnson et al. (2017) have argued, from petrogenetic modelling, that Eoarchaean TTG's were derived from the bases of early basalt sequences with continuous and multiple cycles of volcanism, burial, partial melting and remobilization of TTG's in which there is no a priori need for subduction.

Palaeo- and Mesoarchaean (3.6-2.8 Ga).

Palaeoarchaean rocks occur in all continents and have strikingly similar lithological assemblages and structures that are wholly dissimilar from younger terrains. They form the earliest cratonic cores, such as the Kaapvaal and Pilbara. Today, they are almost earthquake-free, except for mining-induced seismic activity, have low heat-flow, and thick lithosphere, up to 350 km. They have resisted subsequent regional deformation and commonly form hard knots around which younger orogens are moulded. They have, mostly and regionally, low-grade metamorphism and are little eroded; when formed, they were roughly the same crustal thickness as today. Typically, they have the classic TTG dome and greenstone keel structure (Collins et al., 1998; Choukroune et al., 1997). In the Kaapvaal (Anhaeusser et al., 1969) the thick Onverwacht komatiite /chert/greywacke sequence is overlain by the Fig Tree argillite/chert and the Moodies conglomerate/sandstone. The TTG's are in domal structures, typically about 30 km in diameter. The structural fabric generally varies from almost isotropic in the cores of the domes, with increasing flattening fabrics outwards, to plane strain near and at the margins suggesting ballooning. In the greenstones keels, strain is strong from more plane strain fabrics at their margins to vertical stretching fabrics in their cores. Pilbara (Collins et al., 1998), Yellowknife (Drury, 1977) and Dharwar (Choukroune et al., 1997) structures and fabrics are very similar. Bickle et al. (1980) have argued, for the Pilbara, that a pre-doming deformation event was responsible for shortening and early flat-lying structures along the greenstone-granite contacts. We interpret these structures as having formed along the contacts during diapiric inversion. Jackson and Talbot (1989) and Van Kranendonk et al., (2007) have demonstrated the complex polyphase structural complexities that can develop in TTG-greenstone terrains along the margins of mushroom-shaped diapirs – mechanisms include dome margin shear, compression by ballooning, horizontal shearing at the base of overhangs, and compressional constriction in keel cores. Glazner (1994) has shown that mafic intrusions may also enjoy solid-state vertical motions in the continental crust. His model involves the ascent of mafic magma to a position of neutral buoyancy. Upon cooling, the now denser mafic body sinks at rates of several kilometres per million years to a new, deeper, level of neutral buoyancy. These silicic-mafic buoyancy-driven inversions may be important in driving crustal stratification

We interpret these relationships as follows. The Onverwacht (3.56–3.33 Ga) and Fig Tree were laid down as a thick sequence on a hot Eoarchaean TTG basement, followed by crustal inversion, the hot mobilized TTG basement rising as spreading/ballooning/inflating diapiric domes and the greenstones sinking and compressed between the domes (e.g. Schwerdtner et al., 1983). The Moodies was probably mainly restricted to early depressions between the domes and represents the stripping of the Onverwacht, Fig Tree, and some TTG from the dome heads. from the rising domes. These events were terminated by little-deformed, late, high-level, sill-like sheets of potassic granite and small syenite plutons. Although commonly strongly-deformed, the Onverwacht, Fig Tree, and Moodies have a clear stratigraphy in the main Barberton Synclinoriun (Anhaeusser, 1969) that precludes their arrangement in assembled terranes as argued by Lowe (1982). Also, the Onverwacht is clearly a natural sequence of komatiite flows, is not a series of thrust stacks, is not oceanic (see arguments for the Neo-Archaean Belingwe sequence of Zimbabwe below), and cannot be considered to be part of an ophiolite complex as has been suggested (De Wit, 1982; De Wit, 1991). Grosch and Slama (2017) argued for the presence of an ophiolite-type sequence preserved in the Barberton greenstone belt (BGB). They combined new field observations with detrital U-Pb zircon geochronology and geochemistry on fresh drill-core material from the Kromberg type-section sequence of mafic-ultramafic rocks in the Onverwacht Group of the BGB. Trace element geochemistry indicates that the Kromberg metabasalts were derived from the primitive mantle. The εNd values and Nd model ages of the metabasalts record a depleted Archaean mantle source similar to CHUR (chondritic uniform reservoir) with no continental [TTG]) crustal contamination. U-Pb geochronology by laser ablation–ICPMS on detrital zircons from an uppermost chert unit indicate a homogeneous age distribution and a gabbroic source in the greenstone belt, in direct contrast to zircons from felsic conglomerates that structurally underlie the Kromberg sequence. Grosch and Slama (2017) suggested that, collectively, the new data and field observations indicate that the 3.33 Ga Kromberg mafic-ultramafic sequence formed in a juvenile oceanic setting and represents a remnant of tectonically accreted oceanic crust, and that horizontal plate tectonic processes were operating on the Archaean Earth as early as 3.6 Ga. The principal problem in regarding the Kromberg sequence as an ophiolite, generated at an oceanic spreading ridge, is that although it is thrusted (obducted) onto the Noisy Formation diamictites, turbidites and tuffs in the section described, it appears, elsewhere, to be part of the regular Onverwacht volcanic stratigraphy. Also, although there are some lithologies (dunites, gabbros, and pillow basalts) that appear ophiolitic, they are not arranged in anything resembling an ophiolite sequence and there is, critically, no sheeted dyke complex or tectonised peridotite.

The dome and keel structure was generated by vertical inversion of a lighter, hotter Eoarchaean TTG crust overlain by a heavier cooler komatiitic Palaeoarchaean volcanic sequence in a geothermal gradient three times that of today, which would have decreased viscosity and facilitated inversion. The Palaeoarchaean Isua Dome in west Greenland with its flattened and stretched envelope of low-grade meta-volcanics and sediments is very similar to the Barberton TTG domes and was probably also formed during crustal inversion The geology and evolution of Eo- and Palaeoarchaean terrains, in our opinion, is wholly inconsistent with developing in a plate tectonic Earth.

Neoarchaean (2.8 Ga to 2,5 Ga)

By the end of the Mesoarchaean, the Kaapvaal Craton was established upon which the Pongola, Witwatersrand, Ventersdorp, and Transvaal volcanosedimentary sequences were deposited (Fig.1). The Pongola was probably deposited in a rift indicating that the Kaapvaal lithosphere was sufficiently strong to crack and, in Ventersdorp times, to be penetrated by komatiites, basalts and intermediate volcanics. Little-deformed platform carbonates and banded iron formation of the Transvaal Group were deposited across the Kaapvaal Craton as three unconformable sequences in two basins separated by the Vryburg Arch indicating a relatively stable platform and cratonization by this time. Similarly, in Northwestern Australia, the stabilized Pilbara Craton is overlain by the sediments and volcanics of the Fortescue Formation and the banded ironstones of the Hamersley Basin.

However, in the Zimbabwe Craton, the Superior Province and the Yilgarn Craton, the Neoarchaean record is quite different where, within the period 3.1-2.5 Ga, a sub-linear TTG/greenstone crustal fabric was developed. In Zimbabwe, the Belingwe-Bulawayan sequence cycles (2.9 -2.55 Ga) of komatiites, and mafic to silicic volcanics, and sediments arranged in synclinal tracts ranging from broad and open and shallow (Belingwe) to tight and deep (Bulawayo) The distinctive Zimbabwe cover sequence, including thick komatiites, shows minor facies variations but is otherwise uniform and rests with a mapped unconformity upon a Palaeoarchaean TTG-komatiite-basalt (Tokwe/Shibane/Sebakwian) basement (Orpen and Wilson, 1981)) and are intruded by a 2.55 Ga potassic granite suite.. This arrangement is similar to the Palaeoarchaean TTG/greenstone inversion in the Kaapvaal Craton; Zimbabwe well illustrates the various stages of basement- cover inversion from mild and gentle basins to deep and steep keels.

The Superior Province has been interpreted, commonly, as having originated in a plate tectonic regime with volcanic assemblages interpreted as volcanic arcs , sea mounts, oceanic plateaux fore-arcs, and back-arc basins (Capdevila et al., 1982; Corcoran and Mueller, 2007; Davis et al., 1989; Davis et al., 1988; Desrochers et al., 1993; Dimroth et al., 1986; Hoffman, 1990; Jackson and Cruden, 1995; Ludden and Hubert, 1986; Mueller et al., 1996; Percival and Williams, 1989; Polat et al., 1998; Wyman, 1999). The Abitibi, and other, belts in the Superior Province have 2.8-2.7 Ga (Percival and Williams, 1989) tholeiite/komatiite (Pyke et al., 1973), calc-alkaline, and bimodal volcanics (Glikson, 1979; Glikson and Derrick, 1978; Goodwin, 1982; Schwerdtner et al., 1979), with an underlying 3.8-2.8 Ga TTG basement and intruded by 2.7-2.65 "post-orogenic" potassic granites. Commonly, the volcanic sequences are concentric around TTG domes (Goodwin, 1982) giving a regional blobby pattern reminiscent of Pilbara, Dharwar, and Zimbabwe patterns. Volcanic sequences are commonly multi cyclic from mafic to silicic (Goodwin, 1968), unlike stratigraphical arrangements in modern arcs. There are clear belts and some linearity, which Bedard (2018) and Bedard el al. (2013)explain by compression resulting from the convective drag of craton keels causing the collapse of weak zones, sagduction, and the collision of more rigid zones. Calvert et al. (1995) have interpreted shallow-dipping reflection images as evidence of a subduction zone but Ji and Long (2006) argue that other reflectors, such as folded structures can be responsible. Alternatively, the structures could be thrusts developed under the compression inferred by Bedard.

We suggest that both Palaeoarchaean and Neoarchaean terranes were built by the same or very similar TTG basement and basalt-komatiite cover inversion sealed by a terminal craton-sealing high-level granite-syenite event, commonly in thick sills. The Yilgarn Craton has a north-north-west TTG/greenstone belt linearity with TTG domes and greenstone keels and probably originated in the same way. All these Neoarchaean terranes have an older basement upon which komatiite/basalt/calc-alkaline sequences rest, commonly unconformably, and are sealed by a late potassic phase of granite/syenite magmatism. There are no obvious sutures within them. They are arranged in more linear patterns than Palaeoarchaean-style terranes, the first of which was the 3.1 Ga Pietersburg/Murchison Greenstone Belt along the Northern side of the Kapvaal Craton. We regard their volcanic components as pseudo-arcs, containing a significant proportion of andesites but not of subduction-derivation. Lastly, their deep structure is shown in the Kapuskasing Belt (Percival and Card, 1983) along which the Superior Belt was turned up along its north-west margin. Here, almost the whole Superior crust is shown in cross-section as follows from top to base: up to 10 km greenstone belt metavolcanics underlain by 10-20 km of tabular and xenolithic gneissic tonalite and granodiorite, with a basal 20-25 km section of older gneissic granitoid assemblages. This does not favour an interpretation of multiple colliding arcs but rather a volcanic assemblage built upon an older granitoid crust in a stagnant lid.

Many Archaean mafic-ultramafic associations have been claimed to be ophiolites (Friend and Nutman, 2010; Furnes et al., 2014; Furnes et al., 2015; Komiya et al., 1999; Polat et al., 2002) none of which resemble ophiolites. They are scraps and slivers of, variably, pillow lava, gabbro, and serpentinite that could be of any origin; none have sheeted dyke complexes, tectonized harzburgites, or an ordered ophiolite sequence.

Palaeoproterozoic (2.5-1.6 Ga)

During the early Proterozoic, the first arcs appeared in the Birimian (Combs, 2018) from about 2.3 Ga. Narrow linear to curvi-linear deformation zones developed, bounded by cratonic platforms, including the 1.8-1.6 Ga Capricorn Belt, the 2.0 Ga Limpopo and Ubendides of Africa, the 1.9-1.75 Ga Trans-Hudson, Coronation/Wopmay, Cape Smith and Labrador Trough in Canada, and the 1.8-1.7 Ga Mazatzal of the US. The North American orogens weld together seven Archaean provinces and have been interpreted as collision zones (Hoffman et al., 1989). These belts and their margins have many of the hallmarks of Phanerozoic orogenic belts, and were clearly developed in a plate tectonic milieu. They include the characteristic paired positive/negative gravity anomalies, forelands with rifted margin and fore-deep sedimentary sequences, continental shelf prisms, aulacogens, foreland fold and thrust belts, pre- and syn-collisional magmatic zones including adakites and boninites (Wyman, 1999), transcurrent structures, UHT metamorphism and deformation, and substantial basement re-activation. Structures indicative of subduction zones are seen in several places in these belts. This suggests that some form of plate tectonics was operating on Earth from about 2.3 to 1.7 Ga although it is difficult to gauge the degree of plate torsional rigidity. Also there are no blueschists, although thermal conditions may have prevented the formation of low temperature/high pressure metamorphism. The Jormua ophiolite appears not to be a typical supra-subduction zone-ophiolite but rather a slice of an ocean-continent trasition zone obducted about 100 km onto an adjacent platform. Holder et al., (2019) have shown that paired metamorphic belts exist back to the early Proterozoic, but the distinction between the hot, low-pressure and cold high-pressure belts diminishes back in time to zero at 2.5 Ga. The oldest eclogites appear to be outliers to Neo-Proterozoic-Phanerozoic occurrences. The period 2.3-1.7 also includes ophiolites indicative of sea-floor spreading. Although no relics of normal oceanic crust are preserved, except perhaps as accreted fragments, the Jormua ophiolite appears to represent a slice of an ocean-continent transition OCT) zone obducted about 100 km onto an adjacent platform and has characteristics of modern OCT terranes such as Galicia.

In the Palaeoproterozoic, a global mafic outburst is witnessed by a number of LIPS with dykes, sills, and plutons, for example the 2.055 Ga Bushveld Complex, the 2.45 Ga Great Dyke, the 2.45 Ga Matachewan dyke swarm (Hoffman, 1990) and the 2.2 Nipissing mafic dyke and sill complex, all evidence for vigorous plume activity accompanied by extensional deformation in a brittle lithosphere. This vigorous mafic magmatism and a widespread coherent oceanic lithosphere may have been the harbinger and cause of the development of plate tectonics. At 2.5 Ga, there was a massive high volatile flux from the mantle at at least ten times the present rate (Marty et al., 2019) based upon an Archaean ^{129}xenon deficiency relative to modern compositions. Marty et al. argue that this massive outgassing could not have occurred in a plate tectonic regime but represents a 300 million year burst of mantle activity. Also there was a global outbreak in sanukitoid (high Mg granitoids) intrusion at 2.5 Ga. We suggest that, from 2.5 Ga, there began a phase of vigorous convection pluming, mafic magmatism, and lithospheric fragmentation that initiated true subduction and the establishment of plate tectonics, which has continued unabated until today.

The (so-called) Boring "Billion" (1.8-0.8 Ga)

A stumbling block has been argued against plate tectonics during the period 1.8-0.8 Ga, the so-called boring billion. This was the period of the Columbia super-continent leading to the Rodinia supercontinent, which began to break up and disperse during the Neoproterozoic. Stern (2005) suggested that the boring billion was an interval of stagnant lid with widespread potassic magmatism and minor rifting in US granite/rhyolite terrains expressed in Pikes Peak 1.08, Tishomingo 1.375, Wolf River 1.468, and the Wausau Syenite at 1.45-1.45 Ga and 1.34-1.08 Ga (Bickford and Van Schmus, 1985; Bickford et al., 1981). A stagnant lid is difficult to reconcile with the horizontal extension in the Keneewenan Rift, the Grenville, Namaqualand-Natal, and Sveco-Norwegian Orogenies at about 1.0 Ga, and the Gothian (1.5-1.4 Ga) and Hallandian (1.5-1.4 Ga) Orogenies (Stephens, 2020). There are also reasonably well-preserved, likely subduction initiation, ophiolites of about 1.0 Ga age in the East Sayan terrane in southern Siberia (e.g. Belyaev et al., 2017) and in the Yangtse Craton in central China (e.g. Peng et al., 2012). Thus, we see no reasons to deny plate tectonics during the boring billion.

Igneous rock types

Igneous rocks potentially provide the 'smoking gun' needed to establish the existence of plate tectonic processes in the Archean. However, mid-ocean ridge basalts (MORB), as indicators of divergent plate margins, generally owe their compositions to shallow melting of slightly depleted mantle with moderate potential temperatures

(Mckenzie and Bickle, 1988). As these conditions would likely not apply to an ocean ridge in an older, and hence hotter, Earth, the absence of clear examples of Archean MORB (e.g. Pearce, 2008) is insufficient reason to rule out plate tectonics. Much more robust indicators, therefore, are rock types related to convergent plate margins, where subduction of cooler crustal materials beneath hotter mantle should give distinctive compositions that are significantly independent of the age of the Earth at the time. These volcanic arc rock types group as: 1) the products of crystallization of water-rich magmas, such as the calc-alkaline basalt-andesite-dacite-rhyolite (BADR) series, 2) the products of melting of hydrated, depleted mantle such as the boninite-high-Mg andesite (HMA) series, and 3) the products of subducted slab melting such as adakites. Their respective plutonic equivalents include granodiorites, gabbro-norites and TTGs. We examine these three groups in turn below.

Arc-like BADR sequences are rare but do exist in the Archean. Perhaps the best examples are the c. 2.7 Ga Blake River and Confederation Assemblages of the Superior craton (Wyman and Hollings, 2006), part of the c. 2.8-2.7 Ga Youanmi terrane of the western Yilgarn craton (Wyman and Kerrich, 2012) and the 3.12 Ga Whundo Group on the western edge of the Pilbara Craton (Smithies et al., 2005b). They differ in detail from Phanerozic arc assemblages, in particular by being less porphyritic and having a complex range of rock types in addition to the BADR series which may include boninites, high-Mg andesites (HMA), and Nb-enriched basalts (NEB), picrites and adakites. The principal proponents of Archean plate tectonics have attributed this difference to the prevalence of hot and relatively flat subduction in the Archean (Polat and Kerrich, 2006), noting that these rock types are found on present day Earth in areas with hot slab-mantle interfaces such as ridge subduction and slab windows. They also highlight the fact that these rock types do typically have the chemical signatures of subduction such as negative Nb anomalies.

Arguments against a subduction origin for Archean BADR series were summarized by Bédard et al. (2013). A particular question raised by Bédard and others is why such BADR series are so rare in the Archean and why they are typically bimodal (basic-acid), with andesites relatively rare. To paraphrase, and extend their argument, surely 1.5 Ga of hot subduction, with slab fusion adding silica to mantle wedges and abundant water (from subducted serpentinite in particular) available to promote calc-alkaline fractionation, should have produced many more BADR sequences and, in particular, much more andesite. In response to geochemical arguments, they point to evidence that the negative Nb anomalies and related characteristics may be generated by magma-crust interactions as well as subduction (e.g. Pearce, 2008). They propose that Archean BADR sequences are the product of interaction between hot, plume-derived basic magma and Archean crust rather than plate subduction. While accepting the present lack of consensus, we favour the conclusion of Smithies et al. (2018), based on detailed interpretation of the Nb anomalies, that most Archean BADR series are indeed the products of magma-crust interaction but that a small subset do involve some form of transient subduction or sagduction process not necessarily, linked to global plate tectonics.

Of the high-Si, high-Mg rock types, boninites are one of the principal rock types to have been linked to subduction. Boninites were once regarded as diagnostic of subduction, but the discovery of boninites in the Cretaceous Manihiki oceanic plateau (Golowin et al., 2017; Timm et al., 2011), in particular, demonstrated that boninites can also form in plume terranes. It is therefore useful to subdivide past boninite lavas into two groups: 'high-Si boninite, HSB' (with SiO_2> 57 at MgO =8 wt. %); and 'low-Si boninites, LSB' (with SiO_2=52-57 at MgO =8 wt. (Pearce and Reagan, 2019). The HSB group is characteristic of the boninite type area (the Bonin forearc south of Japan), has been found in many ophiolites and basal arc sequences in Phanerozoic and Proterozoic orogenic belts, is characteristic of subduction initiation terranes, and is subduction-specific. The oldest example found so far on Earth marks the start of arc volcanism in the Trans-Hudson Belt at 1.9 Ga (Wyman, 1999). The LSB group co-exists with the HSB group in subduction initiation terranes but is found also at slab edges and in a few oceanic arcs, forearc basins and back-arc basins, and in the Manihiki intraplate setting. LSB have been reported from several Archean settings.

In the detailed evaluations of proposed Archean boninites by Smithies et al. (2004) and Pearce and Reagan (2019), three sets of boninite localities emerged as most likely to be linked to subduction, all classifying as LSB and all having subduction-like geochemical signatures. The first set (the 'Whundo type' of Smithies et al, 2004) comprises the boninites from within the BADR sequences of the Yilgarn (Lowrey et al., 2019), Superior (e.g. Boily and Dion, 2002) and Pilbara (Smithies, 2002) cratons. These are the most likely to have had an origin in an active volcanic arc, and the same arguments for and against plate subduction made in the above discussion of BADR series apply to these boninites.

The second set of boninites (the 'Mallina type') are associated with silicic high-Mg basalts (SHMB) and their plutonic equivalents (high-Mg diorites, or sanukitoids) and are found within intracratonic rifted basins, notably the c. 3.0 Ga Mallina Basin in the Pilbara terrane) (Smithies, 2002; Sun et al., 1988) and large igneous provinces (such as that hosting the 2.7 Ga Stillwater complex, Helz, 1985). These boninite-SHMB associations are restricted to the period 3.0 to 2.0 Ga (Pearce and Reagan, 2019). As the authors cited above have all shown, high-Si, low-Mg compositions are indicative of depleted mantle lithosphere refertilized by a subduction-like component. Thus, although their geological settings show that they are not directly subduction-related, they do carry an inherited record of older subduction events. Their depleted sources and restricted time period are most consistent with a subduction-like refertilization event related to craton accretion. The final set of boninites is that reported for some of the oldest volcanic series, within the small, highly deformed and metamorphosed 3.8-3.7 Ga volcanic sequences at Isua, SW Greenland (Polat et al., 2002) and Nuvvuagittuq, Northern Quebec (O'Neil et al., 2011; Turner et al., 2014). Pearce and Reagan (2019) found that the Isua 'boninite like-rocks' were most similar to the low-Ti basalts – boninite rocks from the Manihiki plateau, but that the Middle Unit of the Nuvvuagittuq sequence did appear to have robust characteristics of low-Si boninites. More work is needed to confirm these inferences and link to plate tectonic processes.

Adakite is another rock type that was once viewed as a subduction indicator, but now has alternative interpretations. Adakites, and their supposed plutonic equivalents, TTGs, have long been explained in terms of subduction, specifically melting of subducted oceanic crust (e.g. Martin et al., 2005). There is no shortage of TTGs in the Archean with the less-differentiated members becoming deeper in origin (residual amphibole to residual garnet), more magnesian and less silicic from the earliest to latest Archean. This is explained by Martin et al. (2005) in the context of a plate tectonic Earth in terms of increasing subduction dip, in which the high-Si adakites (HSA: >3Ga) are silicic slab melts that have ascended directly to the surface, while low-Si adakites (LSA: 3.0-2.5Ga) are slab melts that have reacted with, or instigated melting of, a peridotite mantle wedge before reaching the crust. It is now, however, clear that these rocks could be the product of crustal melting (of plume-derived basalts), in which case the changing characteristics could simply be the result of increasing depth of melting in a steepening crustal geotherm with time. There are many arguments for why crustal melting is the more viable interpretation, the ability of slab melting to generate the observed volume of TTGs being one (Bédard, 2013). Nonetheless, the much less voluminous adakite volcanic rocks within BADR sequences could derive directly from either slab melting, crustal melting or fractional crystallization involving amphibole and/or garnet, as is the case with recent arc-related adakites (e.g. Castillo, 2006).

Overall, therefore, the direct evidence from igneous rock types for plate tectonics is small, with just a handful of localities having rocks indicative of active subduction or sagduction and no evidence for plate tectonics on a global scale. These localities comprise a small number of BADR series that can better be explained by subduction than crustal assimilation together with a comparable number of intraplate boninite-SHMB series that require refertilization of depleted cratonic lithosphere by subduction-like components prior to a second melting event. In our view, as with (Bedard, 2006; Bedard et al., 2013) and Smithies et al. (2018) in particular, and as discussed further in this paper, this can be achieved without global-scale plate tectonics. If geological arguments for stagnant lid Archean tectonics are strong enough, as we believe they are, then processes such as localized craton accretion could explain any subduction signatures so far.

Broad patterns over time

Bradley's (2011) compilation shows that there are a number of compelling changes and trends illustrating gradual and sudden changes in earth history that are likely the results of changes from plume to plate tectonics in a cooling earth. The time distribution of almost everything shows rapid change from about 3.0 to 2.0 Ga, through the late Archaean and early Palaeoproterozoic, with the biggest switch in observed geology at the end of the Archaean Eon. Dome and keel greenstone tectonics dominates the tectonic style of the Archaean, whereas mantled gneiss domes are a minor component of the Proterozoic and younger Earth. The Archaean was a more mafic earth with TTGs and komatiites; komatiite is known after 2.5 Ga only in the Eocene of Gorgona, Columbia. From 3.0-2.0 Ga, there was a steepening rise in the K_2O/Na_2O ratio and decline in MgO and Cr of igneous rocks and a steepening rise of Rb/Sr in sediments. At 2.5 Ga, there was a sharp increase in TiO2, La, Zr, and Sr in igneous rocks.

Mineral deposits with time

Broad trends emerge when relating ore deposit types to geologic age, surficial environment and tectonic setting. The mechanisms by which the crust has evolved, and the amalgamation and dispersal of continents over time, have played critical roles in the formation and preservation of all deposit types.

Figure 2 illustrates the distribution of major ore deposit types as a function of time, and also with respect to the existence of supercontinents. The distribution of ore deposit types over time reveals a broad association with the periodicity of supercontinent assembly and break-up, with fewer deposits forming in periods of supercontinent stasis. Porphyry and epithermal ore forming systems, orogenic Au and MVT Pb-Zn deposits are typically associated with convergence and orogeny (Figure 2) and mineralized districts tend to form close to craton margins and paleo-sutures. Other deposit types may also be linked to dispersal of the supercontinents and divergent margins. It is evident that only diamondiferous kimberlite and PGM-rich layered mafic intrusions have an affinity with intracratonic settings and tend in many cases to form preferentially during periods of supercontinent stasis. The long-lived Mesoproterozoic supercontinent Nuna (or Columbia) is noted for an absence of orogenic ore deposit types, but does contain significant accumulations of continental sediment-hosted metal deposits (SedEx) as well as kimberlites and intrusion-related iron oxide-copper-gold (IOCG) ores.

Evidence exists that the early atmosphere and the precursors to the present oceans formed only at the end of the Hadean era, once the main period of accretion and meteorite bombardment had terminated at circa 3900 Ma (Kasting, 1993). The implications for metallogenesis are that sedimentary and hydrothermal process are likely to have been inconsequential in the Hadean, and any ore deposits that did form at that time were, therefore, largely igneous in character. It is conceivable, for example, that oxide and sulfide mineral segregations accumulated from anorthositic and basaltic magmas at this time. The Eoarchaean, 3800 Ma old, Isua supracrustal belt and associated Itsaq gneisses of western Greenland, for example, comprise mafic and felsic metavolcanics, as well as metasediments, and resembles younger greenstone belts from elsewhere in the world. The Isua belt contains a major chert–magnetite banded iron-formation component as well as minor occurrences of copper–iron sulfides in banded amphibolites and in iron-formations (Appel, 1990). The largest iron-formation contains an estimated 2 billion tons of ore at a grade of 32% Fe. Scheelite mineralization has also been found in both amphibolite and calc–silicate rocks of the Isua belt, an association which suggests a submarine-exhalative origin. The coexistence of banded iron-formations and incipient volcanogenic or sedimentary exhalative, massive sulfide deposits points to circulation of seawater through oceanic crust. Although the zones of known mineralization in the Isua belt are sub-economic, at 3800 years old they represent the oldest known ore deposits on Earth.

Evidence exists through the Meso- and Neoarchaean Eras for substantial crustal amalgamations, such an early Vaalbara continent and later Superia and Sclavia continents (Bleeker, 2003). The existence of Vaalbara (a combination of parts of the Kaapvaal Craton in southern Africa and the Pilbara Craton in Western Australia) receives support from the similarities that exist in the nature and ages of Archaean greenstone belts and supracrustal sequences on the Pilbara and Kaapvaal cratons (Cheney, 1996; Martin et al., 1998), a feature that is especially striking when comparing the Superior-type banded iron-formations of the two regions. It was previously thought that the late Archaean Witwatersrand basin on the Kaapvaal Craton is unique but exploration in Western Australia, has recently revealed the existence of sedimentary sequences of similar age that also appear to be well-endowed with gold mineralization.

The Neoarchaean era represents a period of significant crustal growth and the development of abundant mineralization, formed by processes not unlike those taking place later in Earth history, involving plate subduction, arc magmatism, continent collision and rifting, and cratonic sedimentation. A wide variety of ore-forming processes therefore characterizes this period of Earth history. Well mineralized examples of continental crust that formed in the period 2800–2500 Ma are represented by the granite–greenstone terranes of the Superior Province of Canada, as well as the Yilgarn and Zimbabwe cratons. Greenstone belts formed from arc-related volcanism host important volcanogenic massive sulfide (VMS) Cu–Zn ore bodies, such as those at Kidd Creek and Noranda in the Abitibi greenstone belt of the Superior Province. Off-shore, in more distal environments, chemical sedimentation gave rise to Algoma type banded iron-formations, examples of which include the Adams and Sherman deposits, also in the Abitibi greenstone belt. Greenstone belts formed at this time also comprise komatiitic basalts that, under conditions favorable for magma mixing and contamination, exsolved immiscible Ni–Cu–Fe sulfide fractions to form deposits such as Kambalda in Western Australia and Trojan in Zimbabwe. During and soon after periods of compressive deformation, major suture zones became the focus of hydrothermal fluid flow derived either from metamorphic devolatilization or late-orogenic magmatism. This resulted in the formation of the varied but common styles of orogenic gold mineralization that are typical of most Meso- and Neoarchaean granite–greenstone terranes worldwide. Examples include important deposits such as the Golden Mile in the Kalgoorlie district of Western Australia and the Hollinger–McIntyre deposits of the Abitibi greenstone belt.

Early intracratonic styles of sedimentation, often in foreland basinal settings, gave rise to concentrations of gold and uraninite represented by the ca. 3.0 to 2.7 Ga Witwatersrand basin in South Africa. At least some of this mineralization is placer in origin and was derived by eroding a fertile Archaean hinterland that appears to have been elevated compared to the adjacent basin. The passive margins to these early continents would have developed stable platformal settings onto which laterally extensive Superior type banded iron-formations were deposited. A very significant period for deposition of iron ores such as those of the Hamersley and Transvaal basins of Western Australia and South Africa respectively, as well as the Mesabi range of Minnesota, seems to have been around the Archaean–Proterozoic boundary at 2500 Ma, by which time continental crust was both thick and increasingly rigid.

From a metallogenic perspective, the Palaeoproterozoic is significant because of the major changes that occurred to the atmosphere, especially the rise in atmospheric oxygen levels at around 2300 Ma (GOE). Prior to this, the major oxygen sink was the reduced deep ocean where any photosynthetically produced free oxygen was consumed by the oxidation of volcanic gases, carbon, and ferrous iron. In this environment banded iron-formations, as well as bedded manganese ores, developed, as indicated by the widespread preservation of both Algoma and Superior type iron deposits. The increase in ferric/ferrous iron ratio in the surface environment that accompanied oxyatmoinversion at 2300 Ma, and the accompanying depletion in the soluble iron content of the oceans, resulted in fewer BIFs forming after this time (Bekker et al., 2014; Bekker et al., 2010). The stability of easily oxidizable minerals such as uraninite and pyrite is also to a certain extent dependent on atmospheric oxygen levels and it is, therefore, relevant that major Witwatersrand-type placer deposits did not form after about 2000 Ma.

Metallogenic patterns during the Paleoproterozoic Era were dominated by wide-ranging orogenic processes accompanying plate movements associated with the break-up of Superia and Sclavia and the assembly of Nuna between circa 1800 Ma and 1400 Ma. The break-up of Superia was accompanied, between 2000 and 1700 Ma, by the creation of new oceanic crust and the formation of volcanogenic massive sulfide Cu–Zn deposits such as Flin Flon in Canada, Jerome in Arizona, and the Skellefte (Sweden) – Lokken (Norway) ores of Scandanavia, which are subduction-related in boninite-arc tholeiite sequences.

At 2055 Ma on the Kaapvaal craton, the enormous Bushveld complex, with its world-class PGE, Cr, and Fe–Ti–V reserves, was emplaced, as was the Phalaborwa alkaline complex with its contained Cu–P–Fe–REE mineralization - both generated in intraplate settings (a plume-related Large Igneous Province (LIP) in the case of the Bushveld). In West Africa the period between 2100 Ma and 1900 Ma saw the development of substantial juvenile crust along the margins of the Man Craton during the Eburnean orogeny, accompanied by the formation of extensive orogenic gold mineralization.

The amalgamation of Nuna was followed by a long period of cratonic stability that resulted in the deposition, between 1800 Ma and 1500 Ma, of marginal marine sedimentary basins that host the important SEDEX Pb–Zn ores of eastern Australia (Mount Isa, Broken Hill, and McArthur River) and South Africa (Aggeneys and Gamsberg). Anorogenic magmatism was widespread in Nuna times - in South Australia, for example, the 1590 Ma Roxby Downs granite–rhyolite complex (Johnson and Cross, 1995), host to the enormous magmatic-hydrothermal Olympic Dam iron oxide–Cu–Au-U deposit, was emplaced. Anorogenic granite magmatism at 1880 Ma may also have given rise to the later stages of IOCG style mineralization (such as Estrela, Volp, 2005) in the giant deposits of Carajas, Brazil.

In summary, as more evolved continents formed from late Archaean times and conventional plate tectonic processes appear to have developed, the range of ore deposit types widened, although many have not been preserved because of tectonic reworking and erosion. The Neoarchaean Era, especially from around 2700 Ma, was a period of significant global orogenesis and many ore deposits formed at this time tend to be arc-related and magmatic-hydrothermal to hydrothermal in nature, not unlike those typifying the Phanerozoic Eon when plate tectonic processes were fully extant. The early stages of the Proterozoic Eon were characterized by a number of major crust-forming orogenies, but the period from about 1800 Ma appears to have been marked by longer periods of tectonic quiescence and continental stability. Consequently, although mineral deposits are not unequivocally diagnostic of evolving crust forming processes, it is nevertheless apparent (Figure 2) that a cyclic pattern linked to the development of supercontinents becomes more readily apparent during the Archaean-Proterozoic transition, and it is from this interval of time, therefore, that a conventional style of plate tectonics is likely to have developed.

Diamonds

A key question, is whether any biogenic surface carbon has been taken into the mantle to depths over 150 km to make diamond and if so how this relates to tectonic processes. Diamonds and their inclusions are the only natural samples brought to the surface from depths of >150 km. Most common lithospheric inclusions can be subdivided into the eclogitic (e-type) and peridotitic (p-type) with garnet, clinopyroxene and sulphides being common for both sets. Less commonly, olivine and orthopyroxene are reported in p-type inclusions. Because of the very small amount of impurities, it is impossible to date diamonds; however, diamond inclusions make plausible material for dating. Over the last few decades a number of studies have been done on dating diamond inclusions. Shirey and Richardson (2011) compiled the silicate- and sulphide dates from garnet (bulk), clinopyroxene (bulk) and sulphide (single grain) inclusions worldwide and showed that the oldest inclusions are peridotitic (up to 3.5 Ga), while eclogitic inclusions are younger than 3 Ga. Based on this finding, they concluded that the onset of plate tectonics should have occurred after 3 Ga.

This was contested by two recent studies on the other sets of diamonds and their inclusions. Smart et al.. (2016) analysed carbon and nitrogen isotopic compositions of Archaean placer diamonds from the Kaapvaal Craton which formed between 3.1 and 3.5 Ga, and concluded that diamonds must have crystallised from a source previously exposed to the surface, arguing the onset of subduction and plate tectonics between 3.1 and 3.5 Ga. Smit et al., (2019) investigated mass-independent fractionation (MIF) of sulphur isotopes, recorded prior to 2.4 Ga (Farquhar et al., 2000). Because of the change in atmospheric oxygen between 2.4 and 2.09 Ga, the style of sulphur isotope fractionation changed and only mass-dependent fractionation was recorded after 2.09 Ga. Smit et al., (2019) analysed sulphur isotopes in sulphide inclusions in diamonds from the West Africa, Zimbabwe, Kaapvaal and Slave cratons and reported that MIF sulphur was not observed in the oldest, 3.5 Ga sulphide inclusions from the Slave craton. Younger (<3 Ga) diamond inclusions from the other studied cratons, however, contained MIF sulphur, consistent with the hypothesis that subduction operated from around 3 Ga. Despite very different approaches, all studies on diamonds and diamond inclusions agree at approximately 3-3.1 Ga age for the onset of subduction.

We suggest that diamonds and their inclusions cannot provide a definitive answer to the time of onset of plate tectonics. Sample bias is a serious hurdle to overcome. Dating diamond inclusions (usually very small in size) is not a trivial task, while the lack of samples from a wide range of locations and compositions may constrain obtaining representative ages. Thus, it is possible that eclogitic inclusions older than 3 Ga exist, but have not been sampled yet. The carbon and nitrogen isotopic compositions of diamonds themselves also cannot provide a definitive answer about their source. Carbon isotopic compositions of mantle diamonds have been extensively used to discriminate their origin. Among eclogitic and peridotitic diamond inclusions, the peridotites are significantly more uniform isotopically (with the mean in $\delta^{13}C$ = ~5‰) and are considered to have formed from mantle carbon, while the eclogites show a wide range of carbon isotopic values ($\delta^{13}C$ = -41 to +5‰) attributed to a possible subduction origin. The organic carbon formed at the surface is isotopically much lighter ($\delta^{13}C$ = -40 to 0‰) (Cartigny et al., 2014).

The wide range of carbon isotopic compositions in eclogitic diamonds, does not necessarily reflect a subduction origin (e.g. Cartigny et al., 1998a; Cartigny et al., 1998b). First, to yield such extremely light isotopic signatures ($\delta^{13}C$ =< -25‰) the subducted material would need to consist of only organic carbon, with no input from surface carbonates that dominate surface sediments (~80%) and are isotopically heavier ($\delta^{13}C$ = -15 to -5‰). Second, it is likely that eclogitic and peridotitic diamonds are produced by different mechanisms and involve different fluid speciation. For instance, it has been shown that, even at high temperatures, significant isotopic fractionation could occur between the oxidised (CO_2) and the reduced (CH_4) forms of carbon by means of Rayleigh distillation (Javoy, 1972). This could lead to the light $\delta^{13}C$ in diamonds being precipitated from the methane-rich fluids. Methane-related diamond crystallisation in the Earth's mantle was also detected by combined carbon-nitrogen isotopic studies that do not support a recycled origin of carbon (Cartigny et al., 1998b; Javoy, 1972). Mikhail et al. (2014) have also shown that diamonds in equilibrium with iron carbide are isotopically lighter because of the high carbon isotopic fractionation during the interaction between mantle carbon and native iron at very reduced conditions. Thus, although the surface origin of carbon cannot be ruled out, there are a number of mantle processes that cause carbon isotopic fractionation, triggering the formation of isotopically light diamonds.

The lithosphere

Lithospheric density, thickness and strength, are likely to have increased in response to diminishing asthenospheric temperature diminishes during planetary history. Cooling increases in a plate tectonic planet as slabs are injected deep into the hot mantle, ultimately causing the asthenosphere to cool, thin, stiffen and, eventually to exterminate plate tectonics. Intra-plate alkalic basalts will increase in proportion. The early mantle was hotter by 100-200°C; the lithosphere was, therefore, thinner, weaker and more buoyant, impeding onset of plate tectonics. Lithospheric strength is also essential for plate tectonics to take place; oceanic lithosphere must be strong enough to remain torsionally coherent during plate motion and subduction. Lithospheric strength increases with thickness, so lithospheric density and strength have increased together as the Earth has aged and cooled. There are three aspects of lithospheric weakness that are required for subduction and plate tectonics to happen. First, the lithosphere-asthenosphere boundary must be weak enough for the lithospheric plates to move over it, accomplished by the hotter and weaker asthenosphere and aided by the concentration of volatiles at the interface. Such a weak zone is likely to have existed since lithosphere first formed, very early in Earth history. The second essential weakness required for establishing self-sustaining and asymmetric (one-sided) subduction is evident in the two-dimensional (2-D) numerical experiments of Gerya et al. (2008) which show that the stability, intensity, and mode of subduction require a zone of weak hydrated rocks above the subducted slab. The weak interface is maintained by the release of fluids from the subducted sediments, oceanic crust, and serpentinized upper mantle as the slab sinks and is pressurized and heated. The third weakness required for subduction and plate tectonics is for the development of large-scale, laterally extensive (~1000 km long) weakness through the lithosphere where a subduction zone can nucleate. Without such weakness, subduction and plate tectonics cannot occur. Such trans-lithospheric weaknesses are produced continuously today on Earth by plate boundaries but it is less obvious how the first trans-lithospheric weakness was produced on a stagnant-lid Earth. It seems likely that the first one may have been produced by interaction of a large mantle plume head with old “oceanic” lithosphere (Gerya et al., 2015), during the global outbreak of mafic magmatism between 2.5 and 2.0 Ga.

Subduction and subductability

Plate tectonics requires lithosphere that is dense enough to sink under its own weight. It must be strong enough to remain coherent during self-sustaining subduction, and weak enough to form localized plate boundaries. Such conditions are only likely to happen in a mature silicate planet, as Earth is today. Special circumstances are required for plate tectonics; the planet must be dominantly silicate and conditions must be appropriate for self-sustaining subduction to occur. Because global plate motions are mostly powered by the sinking of the negatively buoyant oceanic lithosphere in subduction zones (although significant roles for ridge push and mantle convection drag for modifying these motions has been repeatedly proposed) the lithosphere must be denser than the underlying asthenosphere.

Self-sustaining subduction of the oceanic lithosphere, as the engine of plate tectonics, cannot occur on a constant-size Earth without spreading ridges to balance the loss of area by subduction. Therefore, in a stagnant lid planet (>2,5 Ga), MORB cannot be erupted and the non-continental crust must be formed in some other way, probably by the vertical accumulation of basalts, komatiites, and minor differentiated silicics. Because of the higher temperature of the asthenosphere, the lithosphere would have been thinner and the mafic-ultramafic crust thicker. It seems unlikely that the pre-2.5 Ga crust was wholly continental because subduction would have been unable to start. Also, the present crustal thickness of Archaean cratons was achieved immediately at the end of crustal inversion and the intrusion of late potassic granites and sanukitoids. Most of Earth's continental crust was generated by 2.5 Ga. possibly by the end of the Eoarchaean. Had this crust been globally enveloping, some two thirds of this continental crust would have to have been lost by some process such as tectonic decretion by subduction erosion after 2.5 Ga. Subduction, and hence plate tectonics in a silicate planet, needs lithosphere that is sufficiently dense, stiff, and coherent to sink at weak zones rather than float. The integrated density of present-day oceanic lithosphere younger than about 20 million years will float unless attached to older lithosphere, whereas lithospere older than 20 million years is unstable and can sink. Stern calls this the crossover time, when the integrated density of the lithosphere equals the density of the asthenosphere (3.25). A young stagnant lid of non-continental lithosphere will be more buoyant than young MORB lithosphere because of its thick komatiitic crust and its thinner lithosphere resulting from a hotter asthenosphere. A stagnant lid of oceanic lithosphere above a hotter asthenosphere that constantly thermally eroded the lid would not progressively thicken as does lithosphere moving away from a ridge. Therefore, the crossover time of a stagnant lid lithosphere would have been correspondingly longer and preventing the onset of subduction and plate tectonics. We see no evidence of plate tectonics until after 2.5 Ga; therefore, the Archaean crossover time was probably very long, subduction was not possible, and plate tectonics was delayed until after 2.5 Ga.

A stagnant lid is the default mode of planetary tectonics (Stern et al., 2018; Stern et al., 2016). It is the potential energy resulting from denser lithosphere on above` weak asthenosphere that provides most of the energy for plate motion (Forsyth and Uyeda, 1975; Lithgow-Bertelloni, 2014). Such a density inversion does not exist for continental lithosphere or for oceanic plateaus, where low-density crust reduces the overall lithospheric density. Such a density inversion was less likely early in Earth history, when hotter mantle resulted in thinner mantle lithosphere and thicker oceanic crust.

Continental crust formation and destruction

One of the unique features of Earth is its two types of crust: 40 km thick dioritic continental crust and 6 km thick basaltic oceanic crust. Human beings only exist because there are continents for us to evolve on and exploit. If plate tectonics has been extant since 4.6 Ga, Earth's continental crust must be growing with time and must be a product of plate tectonics. On the other hand, if plate tectonics began between 2.5 and 2.3 Ga, there must have been another mechanism for forming continental crust because there are many crustal tracts older than this.

The present-day flux from mantle to crust is basaltic and yet the average composition of the continental crust is andesitic. This is the crust composition paradox. A new solution to this paradox is proposed whereby the secular evolution in the composition of the continental crust reflects a changing flux from mantle to crust over time. Thus it is proposed that the present-day composition of the continental crust is a time-integrated average. Crustal growth curves show that at least 50% of the continental crust had formed by the end of the Archaean (Dhuime et al., 2012, 2015). A mass balance model based upon a tonalite-trondhjemite-granodiorite (TTG) composition compositional model for the Archaean continental crust shows that the post-Archaean mantle to crust flux was predominantly basaltic and likely a mix of arc-plume basalts. Trace element modeling, however, reveals that additional processes also contributed to the average crust composition. Balancing Y, Ho, and Yb concentrations requires a garnetiferous mafic granulite composition for the lower Archaean crust, which in turn drives the post-Archaean flux toward a high-magnesium andesite. This suggests that there was a slab melt contribution to the continents, in addition to basalt. An excess of fluid mobile elements in the continental crust can be explained either by the addition of a slab melt or small fraction melts. A deficiency in Sr requires that the post-Archaean crustal composition has been modified by erosion. Both Archaean and post-Archaean continental crust contain contributions from basalt and a slab melt. In the Archaean crust the slab melt contribution is dominant. In the post-Archaean crust the basaltic contribution is dominant.

There remains, however, the problem of the global distribution of Archaean crust. Some, possibly a substantial amount (difficult to ascertain), was recycled as sediment, melted or partially melted to form younger granitoid plutons and silicic/intermediate volcanics, and structurally reworked into younger orogens. Possible geometric solutions are;

1. There was a global continental coverage of Archaean crust that was segmented and shortened into the extant Archaean cratons, which cannot be correct because there is insufficient post-Archaean shortening in the Archaean cratons.

2. During the later plate tectonic regime, large amounts of a global Archaean continental crust were lost by tectonic decretion/subduction erosion. Tectonic decretion is sufficient, today, that crustal subtraction is slightly larger than addition (Dewey and Windley, 1981). Armstrong (1968) argued that the volume of the continental crust has remained roughly constant since about 3.8 Ga, either by a subtraction addition balance or that all crustal growth was during the Archaean. Fyfe (1978) went further, arguing for continental growth until about 2.3 Ga then continental diminution to the present day.

3. The present volume of continental crust in the present cratonic nuclei is all that ever existed. This means that about 86% of Earth, during the Archaean, was covered with remnant Hadean oceanic lithosphere, none of which is preserved directly or its existence suggested by rocks generated by plate tectonics. We think it inconceivable that a non-plate tectonic inversion process could have been occurring to generate the structure of the continental crust, while, in a surrounding ocean, plate tectonics was generating ridges, arcs and collisions from which none of the rock results are preserved. Therefore, we incline towards massive tectonic decretion from about 2.3 Ga and subduction of any remaining Hadean lithosphere.

Evolution

Biological evolution is driven by isolation and competition. Isolation allows new species to evolve and competition selects the organism that has best adapted to the environment. Plate tectonics is an mechanism for creating and destroying biological environments such as continents, continental shelves, deep ocean basins, island arcs, and mountains, and therefore is also an unparalleled engine for isolating and recombining ecosystems, favouring speciation and competition. In contrast, stagnant lid tectonics has far less ability to create and destroy biological environments. The implications for evolution are obvious: plate tectonics favours rapid evolution, stagnant lid tectonics does not. Yet we see almost three and a half billion years of stromatolites and other primitive organisms with a rapid metazoan evolution in the Ediacaran. Thus, almost two billion years of plate tectonics did not produce a metazoan evolution. Perhaps, in spite of the Great Oxidation event at 2.3 Ga, oxygen levels only became sufficient, when they climbed rapidly from about 3% to 13% at about 0.7 Ga.

Conclusions

We consider 2.5 Ga, the end of the Archaean as the beginning of a time of fundamental change in Earth's tectonic behaviour that led to the establishment of plate tectonics by about 2.3 Ga. Consensus is not a necessary prerequisite for truth but there is general agreement, for all the reasons given by Stern (2005) that plate tectonic has operated since at least 0.7 Ga. These include ophiolites, blueschists, UHP metamorphism, sutures and collision zones, paired metamorphic belts, zones of substantial crustal shortening, major strike-slip faulting, and linear adakite zones. Collectively and from their geological arrangements, these have a clear plate tectonic fingerprint. Palaeomagnetism shows that there was relative motion among continental cratons back to at least 1880 (Condie and Kroener, 2008), The Wopmay and other Palaeoproterozoic orogens such as the Trans-Hudson, the Capricorn, and Limpopo, are belts of, variably, adakites, miogeoclinal platforms, the classic paired positive-negative gravity anomalies, suturing, and horizontal shortening, all of which suggest a plate tectonic origin back to at least about 2.0 Ga. These orogenic belts weld together Archaean cratons and are best-described as collisional orogens, themselves an indication of the former subduction of some form of oceanic tract., either a stagnant lid lithosphere or lithosphere generated by sea-floor spreading; if the latter, this pushes plate tectonics back before 2.0 Ga. The 2.04-2.24 Ga Birimian in the Reguibat of Mauritania and the Man-Leo Shield of Cote d'Ivoire (Combs, 2018) is very similar to the Neoproterozoic Pan-African and models for the Phanerozoic of central Asia (Sengor et al., 1993), comprising huge tracts of accreted volcanic arcs, also takes plate tectonics back to at least 2.3 Ga. Holder et al (2019) made the key analysis and observation that paired metamorphic belts, distinctive of convergent plate boundary zones, become weaker in their bimodality back in time until the bimodality vanishes between 2.2 and 2.4 Ga, supporting the advent of plate tectonics at this time.

Many workers in Archaean (>2.5 Ga) terrains have developed plate tectonic models for these rocks (see Korenaga, 2013)), although there is a small group of dissenters (Davies, 1992; Dewey, 2018a; Dewey, 2018b; Dewey, 2019; Hamilton, 1998; Hamilton, 2003; Hammond and Nisbet, 1992; Padgham, 1992). The blobby to sub-linear TTG dome and greenstone keel pattern that dominates the upper crust from about 3.6-2.6 Ga and the stratigraphical sequences of the Archaean show no patterns that can be related to plate boundary zones, and no lithological assemblages like those of the plate tectonic world. The dome and keel patterns were developed by crustal inversion of a light, hot, TTG crust beneath a colder, heavier, volcanic crust above with TTG domes, containing radionuclides and gold, rising and ballooning and greenstones sinking by sagduction and drip. The greenstone keels are characterised by steep to vertical prolate fabrics while the TTG fabrics change from almost isotropic in dome centres through increasing flattening to plane strain at the margins. Strain patterns are, locally, more complex and polyphase but do not imply development prior to or independent of diapiric ballooning (Snowden and Bickle, 1976). The common TTG gneisses may have been formed by flow and flattening beneath the plutons. Archaean TTGs are distinct from later TTGs in being silica-alumina-soda rich, potash-iron magnesium poor with low heavy REE and fractionated rare earths.

Komatiites, are mainly Archaean; they cannot be oceanic crust because they occur with sediments, basalts and silicic rocks in thick sequences on continental crust; they were likely developed by peridotite partial melting in plumes at mantle potential temperatures at least 200°C higher than present. The sediments, usually water-lain are commonly volcaniclastic but do not show the structural patterns of accretionary prisms. Claims have been made that ultramafic-basalt associations are parts of ophiolite complexes and represent slices of obducted oceanic lithosphere (such as at Isua or in the Kromberg Formation at Barberton; Grosch and Slama, 2017), but although they contain pillow lavas and are therefore sub-aqueous, do not resemble, even remotely, the classic ophiolite sequence of the Phanerozoic, lacking key components such as sheeted dyke complexes and residual tectonized harzburgites. Archaean lithologies, structures, and patterns bear only a passing resemblance to younger ones but never in the same arrangements, and relationships. The Archaean has lithologies, patterns and structures that are unknown or rare in later terrains and contains very few that are characteristic of plate boundary zones.

We emphasise that one cannot use a simple shopping list of features and characteristics either all of or any one which must be observed to establish a plate tectonic origin. Also, no broad rock type name such as andesite, tonalite, or boninite can be used as definitive; these are merely names that conceal wide variations in petrology, geochemistry and significance. It is necessary to specify the precise petrology and chemistry in rock suites that show great variation among a variety of tectonic situations. Particularly egregious is the use of the general name adakite as pejorative proof of a volcanic arc; this is tautology. Scraps of serpentinite, gabbro, and pillow basalt do not necessarily constitute an ophiolite - lithologies must be arranged in sequence with sheeted dykes to have any validity as oceanic crust and mantle generated at a ridge axis. We suggest that there are three types of evidence used to argue for plate tectonics in the Precambrian, especially during the Archaean:

1. Based upon incorrect data.

2. Based upon incorrect interpretations of correct data.

3. Based upon correct data and interpretation.

It is the latter which best describes plate tectonics back to 2.0 Ga and, perhaps to 2.5 Ga. It is the pattern and arrangement of rock types in linear and curvilinear belts that separate older platforms and cratons and are indicative of extinct plate boundary zones, just as for recent and present ones. The difference is that whereas today we can see oceanic lithosphere entering modern subduction zones, old subduction zones can be inferred only from sutures, “adakite” belts, orogens and palaeomagnetism.

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Figure 1. Outline of domains, rock suites, sequences and events referred to in text

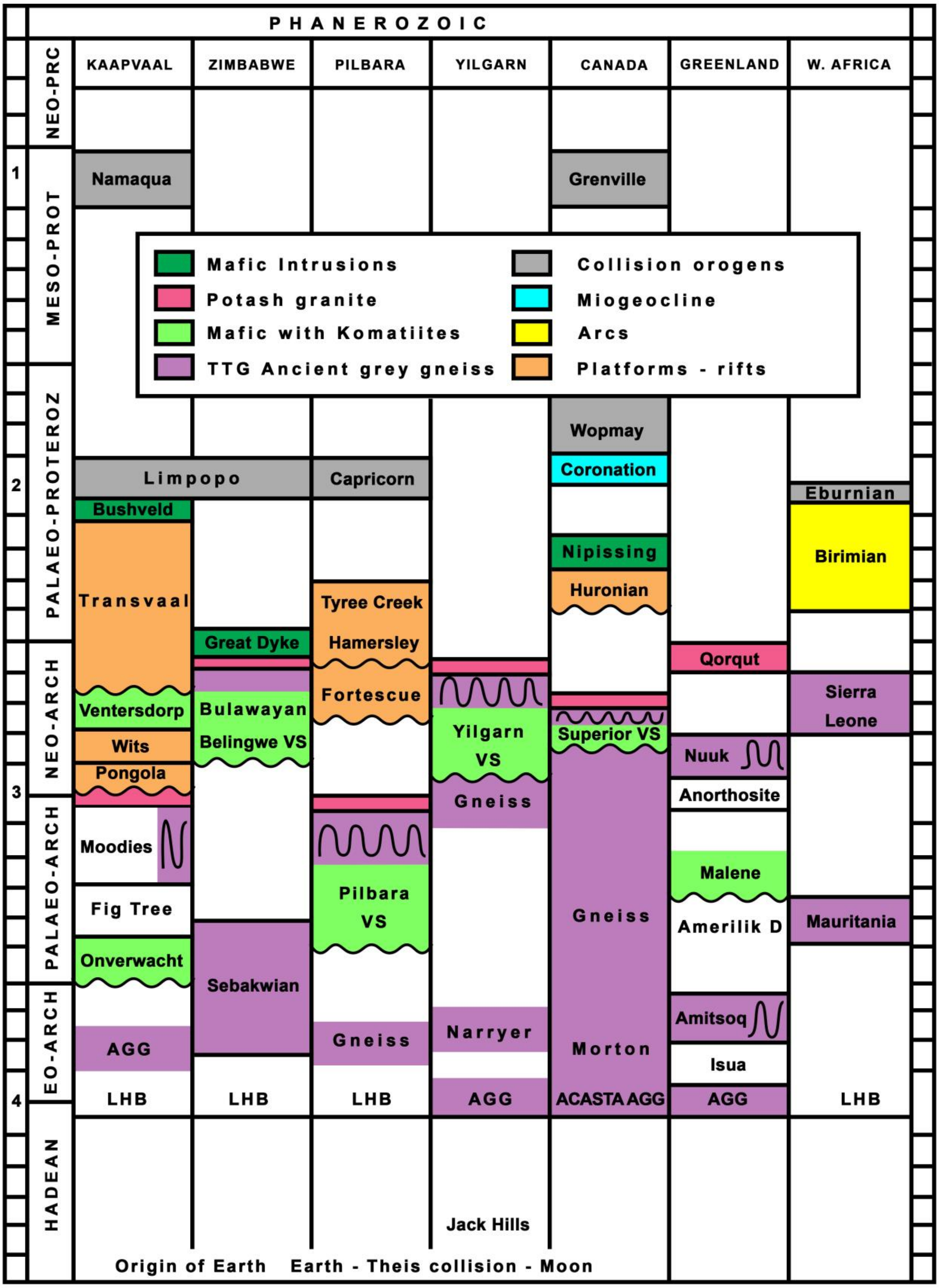

Figure 2: Distribution of major ore deposit types as a function of time and the supercontinent cycle (after Robb, 2020); IOCG – Iron Oxide-Copper-Gold/SEDEX – Sedimentary Exhalative/VMS – Volcanogenic Massive Sulphide/MVT – Mississippi Valley Type